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Hydraulic Geometry of Gravel-Bed Rivers

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HYDRAULIC GEOMETRY OF GRAVEL-BED RIVERS

Final Report

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Abstract

Field measurements of channel and bed material properties are used to develop a network-based model of bed-load transport capacity in Halfmoon Creek, a high-gradient stream in the Rocky Mountains of Colorado. Measurements of channel geometry, slope, and bed sediment texture (surface and substrate) were taken in 27 separate reaches with average gradients ranging from 0.6 to 6%. These data are used to examine interactions among key variables affecting the processes of bed load transport. The analysis shows that the grain-size distribution of the substrate is consistent throughout the basin; however, the distinction between surface and substrate texture varies systematically with channel gradient and shear stress. Reaches with high shear stress are characterized by very coarse surface layers, which have a high threshold for transport, and also limit the mobility of the sediment supplied. In reaches with lower gradient and shear stress, the difference between surface and substrate texture is less, increasing both the frequency of transport and the mobility of the bed load. Modeled estimates of bed-load transport rate and transport frequency indicate that annual sediment yields increase nearly linearly with discharge, suggesting a simple scaling of bed-load transport capacity as discharge and drainage area increase downstream.

Introduction

Headwater streams are frequently described as complex systems that evolve over time in response to a mix of geomorphic processes [*Knighton*, 1998; *Wohl*, 2000]. Hillslope transport of sediment from landslides and debris flows may dominate sediment supply in some watersheds, locally forcing changes in channel gradient and grain size [*Grant et al.*, 1990; *Montgomery and Buffington*, 1997; *Brummer and Montgomery*, 2003]. Elsewhere, fluvial sediment transport dominates and channel morphology reflects the orderly transfer of water and sediment according to established principles [*Emmett*, 1975; *Pizzuto*, 1992; *Whiting et al.*, 1999]. These and many other observations have led to generalized conceptual models and stream classification systems which characterize the headwater reaches of rivers as being primarily erosional, whereas intermediate or distal reaches are considered transitional to depositional [*Schumm*, 1977; *Church*, 1992; *Montgomery and Buffington*, 1997].

Modeling erosion and sediment transport across these different domains is problematic, and perhaps only loosely tied to position within the drainage basin. In headwater streams, the most important conceptual issue concerns the distinction between alluvial channels in which the sediment-transport capacity, Q_c, balances the sediment supply, Q_s, and non-alluvial channels in which $Q_c > Q_s$. There is a tendency to lump mountain streams into the latter category for two reasons: (1) channel gradients are generally high (> 2%), therefore, the shear stress or stream power available to move sediment appears to be quite high [*Montgomery and Buffington*, 1997]; and (2) the sediment seen on the bed (cobbles and gravels) is typically much coarser than the sediment collected during short sampling intervals (sand and granules), and potentially coarser than the material sampled from the substrate (bulk bed material). Observations that the bed load is finer than the substrate [*Church et al., 1991*; *Lisle*, 1995; *Wathen et al.*, 1995; *Whiting et al.*,

1999; *Habersack and Laronne*, 2001; *Ryan and Emmett*, 2002] imply that the load is derived from sources other than the bed, and that transport rates are driven more by sediment supply than by flow conditions [*Dietrich et al.*, 1989; *Buffington and Montgomery*, 1999b]. However, a number of other field studies and data sets indicate that transport in headwater streams is closely coupled to flow properties and bed material characteristics [*Parker et al.*, 1982; *Andrews*, 1984; *Andrews*, 1994; *Lisle*, 1995; *Powell et al.*, 2001; *Torizzo and Pitlick*, 2004]. It appears that the range of conditions in headwater basins is quite large, and it is not evident that bed load transport data, by itself, conveys the information needed in the analysis of watershed sediment supply.

In this paper we propose an alternative approach for assessing sediment-transport capacity in headwater streams. We start with the premise that the volume or mass of bed load carried in alluvial rivers is governed by a handful of inter-related factors, including the water discharge, the channel geometry (width and depth), the slope, and the grain sizes available for transport [*Gilbert*, 1914; *Mackin*, 1948; *Rubey*, 1952; *Henderson*, 1966; *Ferguson*, 1986; *Carson and Griffiths*, 1987; *Howard et al.*, 1994]. None of these factors can be excluded in assessing the sediment-transport capacity of a stream system. It is particularly important to include grain size as a variable, due to its importance in scaling the transport rate, and in defining thresholds for motion. The distinction between surface and substrate (or subsurface) grain sizes is of further importance in gravel-bed rivers because local coarsening of the bed surface limits the mobility of finer sizes that would otherwise move quite easily [*Parker*, 1990]. Modifications of bed surface textures provide a mechanism for channels to adjust to changes in transport capacity and sediment supply [*Dietrich et al.*, 1989; *Lisle et al.*, 1993; *Buffington and Montgomery*, 1999a,b; *Wilcock*, 2001; *Lisle and Church*, 2002], and, in the absence of major sediment inputs, may result in downstream fining within sedimentary links [*Rice*, 1999] or reaches of rapidly

declining shear stress [*Ferguson et al.*, 1996]. The variables needed to estimate sediment-transport rates (width, depth, slope, grain size) are readily measured through a channel network, therefore, we propose that their joint variation should reflect changes in sediment-transport capacity. This approach is consistent with transport-based theories for hydraulic geometry [*Parker*, 1979; *Pizzuto*, 1992], and takes advantage of the fact that channel properties are not nearly as variable in space or time as is the process of bed load transport.

Linkages between channel morphology, grain size and sediment-transport capacity are investigated here using measurements from Halfmoon Creek, a headwater stream located in the Sawatch Range of central Colorado. Halfmoon Creek contains a mix of self-formed and forced (non-alluvial) reaches. This setting is typical of mountain streams where channel and sediment characteristics change rapidly over short distances. Our measurements emphasize geomorphic conditions in self-formed reaches with gravel- to cobble-size bed material and average gradients ranging from 0.6-6%. The self-formed reaches are discontinuous, but they occur throughout the watershed, including the lowest-order tributary segments. The forced reaches are associated with glacial steps or moraines situated throughout the basin; these reaches are steep (> 10%), and act effectively as conduits that pass sand- and gravel-sized sediment to adjoining alluvial reaches. Estimates of total transport through the alluvial reaches are computed on the basis of a bed-load transport relation and a regional flow duration curve. These estimates are then used to assess changes in the mass balance of sediment through the stream network. We hypothesize that if the stream system is predominantly supply limited ($Q_c > Q_s$) then the estimated loads will peak near the headwaters and decline rapidly toward the basin outlet. If the stream is transporting at capacity ($Q_c = Q_s$), then we predict that the bed load will increase monotonically toward the basin outlet.

Study Area

Halfmoon Creek is located in the Sawatch Mountains of central Colorado (Fig. 1). The drainage basin ranges in elevation from 4400 to 2990 m, and includes the two highest peaks in Colorado (Mt. Elbert and Mt. Massive). The entire basin has been glaciated. Talus slopes and bedrock exposures of Precambrian gneiss and schist occur throughout the headwaters (> 3500m elevation); the lower basin includes open, U-shaped valleys bordered by recessional and lateral moraines, which are now forested. The setting is typical of alpine basins in Colorado, where Quaternary glaciation has had a primary influence on basin morphology. Annual peak flows in Halfmoon Creek are derived from snowmelt runoff, resulting in a regular inter-annual hydrograph with a mean annual flow of 0.85 m^3/s and a mean annual flood of 8.0 m^3/s as recorded at the USGS gauge near the downstream end of the study area (period of record 1947-2002). Summer thunderstorms are common but they are of short duration and typically do not produce much runoff. At the furthest downstream site (Fig. 1) the creek drains an area of 64 km^2 and the flow is unregulated.

Stream sediment is derived from localized bank erosion and mass wasting of hillslopes, glacial moraines and debris fans. Overall, the basin is relatively stable against erosion, and the prior effects of glaciation limit the coupling between hillslope processes and channel morphology in a number of places [*Church*, 2002]. Nevertheless, there is trace evidence of small hillslope failures resulting from a localized storm in the center of the basin, and evidence of active sediment transport along most of the main channel. Bed load sediment yields have not been measured, but deposits left in a retention basin at the outlet of the basin suggest that the annual sediment yield is on the order of 10 m^3/yr.

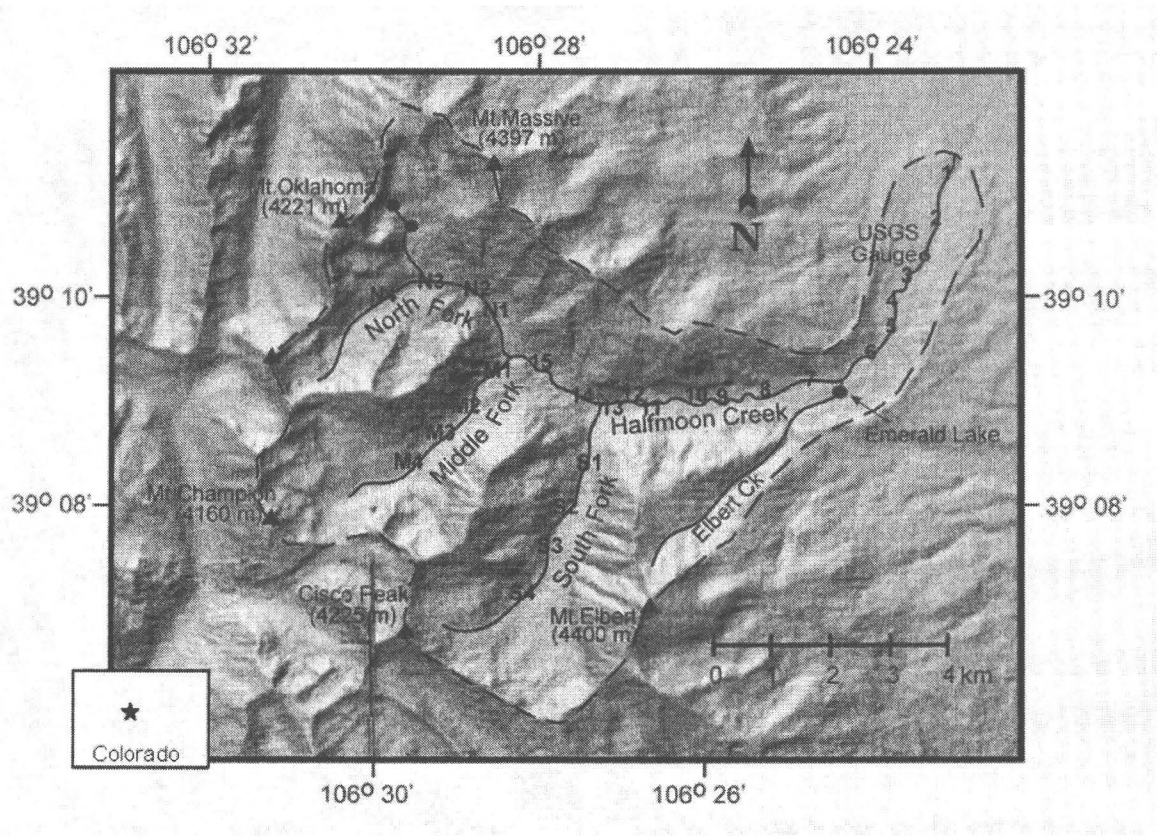

Figure 1. Location map of study sites within the Halfmoon Creek basin generated from 30m DEMs. X's represent the 27 study reaches.

Measurements were taken in 27 alluvial reaches along the main channel and in three tributary valleys, herein differentiated as South, Middle, and North Halfmoon Creeks (Fig. 1). The study sites are located at closely spaced intervals (every 0.5-1.5 km) along the main-stem and tributary segments. Site characteristics vary considerably (Table 1). In general, the study reaches can be discriminated between those occurring in hanging valleys, which are separated from the main valley by cascade-type channels at glacial steps, and those occurring in the main valley itself (Fig. 2). Although many of the reaches are bordered by forest, woody debris is uncommon in the channel.

Table 1. Study site characteristics. *Dist.* is the distance downstream, A is the drainage area, $\hat{Q}_b$ is the estimated bankfull discharge, S is the slope, D_{50} is the median grain size of the surface bed material, D_{50s} is the median grain size of the subsurface bed material, and R is the bankfull hydraulic radius, B is the bankfull width, τ is the bankfull shear stress, τ^* is the bankfull dimensionless shear stress, and Q_s is the modeled annual sediment yield.

Site	Dist. (km)	A (km^2)	Q_b (m^3/s)	S	D_{50} (m)	D_{50s} (m)	R (m)	B (m)	τ (N/m^2)	τ^*	Q_s (t/yr)
1	14.6	64.0	9.5	0.0180	0.072	0.036	0.49	10.4	87	0.075	172
2	14.2	63.7	7.3	0.0140	0.061	0.023	0.46	10.8	63	0.064	171
3	13.3	62.5	6.4	0.0084	0.050	0.018	0.47	9.7	39	0.048	171
4	12.3	61.7	6.8	0.0085	0.052	0.019	0.58	7.8	48	0.057	144
5	11.7	61.2	5.5	0.0086	0.055	0.020	0.49	8.0	41	0.046	160
6	10.6	59.7	6.5	0.0120	0.062	0.019	0.48	9.3	56	0.056	170
7	9.3	54.2	6.5	0.0200	0.078	0.033	0.52	7.6	102	0.081	131
8	8.5	53.0	5.0	0.0064	0.030	0.019	0.37	10.7	23	0.048	107
9	7.5	51.0	7.1	0.0097	0.038	0.019	0.42	10.4	40	0.065	108
10	6.7	49.1	8.4	0.0130	0.053	0.018	0.45	11.8	57	0.067	159
11	6.4	48.6	7.8	0.0160	0.066	0.023	0.44	10.9	69	0.065	174
12	6.0	47.3	8.3	0.0220	0.068	0.018	0.50	8.3	108	0.097	109
13	5.6	46.1	8.0	0.0230	0.072	0.027	0.48	8.4	108	0.093	117
14	5.3	30.3	7.0	0.0370	0.086	N.A.	0.47	6.9	170	0.122	96.5
15	4.6	28.4	5.7	0.0440	0.078	0.030	0.41	7.0	177	0.140	79.8
N1	2.5	13.4	2.6	0.0450	0.081	0.036	0.35	4.5	154	0.118	54.3
N2	2.1	12.1	3.5	0.0220	0.066	0.019	0.44	3.8	95	0.089	44.2
N3	1.2	9.8	1.7	0.0230	0.066	0.026	0.31	3.8	70	0.066	47.1
N4	0.8	4.6	1.5	0.0460	0.055	0.034	0.26	3.6	117	0.133	24.2
M1	3.3	11.1	2.1	0.0400	0.060	0.020	0.32	3.9	125	0.129	30.9
M2	1.8	9.3	2.5	0.0360	0.066	0.020	0.28	5.5	99	0.092	52.7
M3	0.9	6.7	1.6	0.0320	0.065	0.017	0.29	4.3	91	0.087	42.6
M4	0.0	3.2	1.0	0.0470	0.060	0.020	0.29	2.5	134	0.138	18.9
S1	4.2	13.8	2.3	0.0410	0.069	0.019	0.35	4.2	141	0.127	41.1
S2	2.9	10.5	1.4	0.0400	0.071	0.020	0.35	2.5	137	0.119	25.2
S3	1.9	7.0	1.0	0.0200	0.051	0.035	0.28	3.0	55	0.067	27.9
S4	1.4	3.1	0.7	0.0670	0.074	0.026	0.21	2.7	138	0.115	26.9

Twelve of the study sites are located in tributary valleys. These sites are located in plane-bed reaches that are steep, but otherwise, self-formed (Fig. 2a). Fifteen study sites lie along the main channel of Halfmoon Creek. These sites are located in alluvial plane-bed or pool-riffle reaches where gradients are somewhat less (Fig. 2b). Downstream of the confluence with the south fork, the slope of the main channel decreases consistently as the creek flows from steep, forested terrain into a wide alluvial valley through which the stream meanders. The valley then narrows where a small tributary, Elbert Creek, has produced an alluvial fan and a steepening of the valley profile. Downstream of the fan, the valley again opens up, allowing the stream to meander across a floodplain (Fig. 2c). The east end of the valley is bordered by lateral moraines which restrict the channel and steepen the gradient through the two downstream sites.

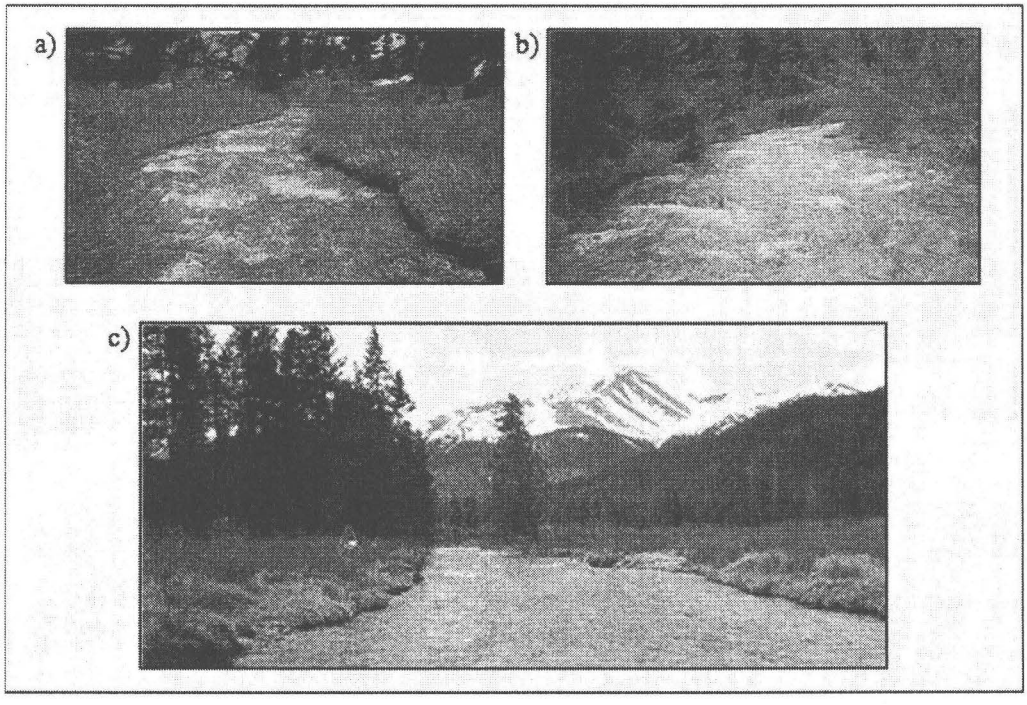

Figure 2. Characteristic examples of study reaches. A) Downstream view of site N3, north fork of Halfmoon Creek. Site is located in a broad glacial valley at an elevation of 3473 m; $S = 0.023$, $B = 3.8$ m. B) Upstream view of site 12, main channel of Halfmoon Creek. Site is located in a forested area at an elevation of 3105 m; $S = 0.022$, $B = 8.3$ m. C) Upstream view of site 3, main channel of Halfmoon Creek. Mt. Elbert in the background. Site is located in an open meadow at an elevation of 3031 m; $S = 0.0084$, $B = 9.7$ m.

Methods

Basin-wide trends in sediment-transport capacity were determined by calculating the annual bed-load sediment yield at each of the 27 study sites. Consistent measurements of channel and bed material properties were made at each site in order to estimate parameters for calculating transport rates (details of the calculations are discussed below). In addition to the field data, a key element of the analysis includes the development of an empirical relation for adjusting the threshold for bed load transport in steep streams.

The study sites were located in relatively straight reaches to limit complications arising from changes in channel curvature and topography. A minimum of three cross-sections was surveyed with a level at each site in order to determine the bankfull channel geometry. Measurements of both the bed and water-surface profile were made over an approximate length of ten times the channel width in order to determine the reach-average slope. Statistical tests of the regression coefficients indicated there was no significant difference in bed and water surface slopes ($p < 0.05$), thus the latter were used to approximate the energy slope. Bankfull channel dimensions were determined by a distinct break in bank slope. Measurements of bankfull depth at the left bank and right bank were generally similar (within 5 cm), with the lower value being used in calculations to define a bankfull hydraulic radius and width (Table 1).

Surface sediment samples were taken at each site using a variation of the *Wolman* [1954] method, with a minimum of 100 particles sampled at random. Particle sizes were measured at 1/2-phi intervals using a metal template (gravelometer). Bulk samples of the subsurface sediment were also obtained at all but one site in order to determine the ratio of surface to subsurface grain-size fractions, degree of bed armoring, sorting, and trends in downstream fining. Bulk samples were large enough to ensure that the largest grain was no more than 5% of

the total weight and usually much less [*Church et al.*, 1987]. The coarse fraction (>32 mm) of the subsurface samples was sieved in the field and the fine fraction (<32 mm) was sieved in the lab, again at 1/2-phi intervals. Graphical plots of grain size versus cumulative frequency were used to determine percentiles (D_{84}, D_{50}, D_{16}) of the size distribution of each sample (Table 1).

Annual bed load sediment yields were calculated for each site following the procedure outlined by *Wolman and Miller* [1960]. At each site we calculated the instantaneous bed load transport rate, Q_s, for a given discharge, Q, and then multiplied that value by the frequency of that discharge, $f(Q)$. Although straightforward in principle, the computations of sediment load and discharge frequency involve a number of assumptions and uncertainties which we discuss below.

The estimates of bed load transport are based on the equation of *Parker* [1979], which can be written in a general form as:

$$q^* = 11.2\frac{(\tau^* - \tau^*_c)^{4.5}}{\tau^{*3}} \qquad (1)$$

where q^* is a dimensionless transport parameter, τ^* is the dimensionless shear stress, and τ^*_c is the critical dimensionless shear stress. An additional term, correcting for the effects of bed forms or other sources of roughness might be included in (1); however, we leave this discussion for later. The dimensionless shear stress is

$$\tau^* = \frac{\tau}{(\rho_s - \rho)gD_{50}}, \qquad (2)$$

where τ is the boundary shear stress, ρ_s and ρ are the densities of sediment and water, respectively, g is the gravitational acceleration, and D_{50} is the median grain size of the bed sediment. The transport parameter can be expanded as follows:

$$q^* = \frac{q_b}{\left[(s-1)gD_{50}^{\ 3}\right]^{0.5}}, \tag{3}$$

where q_b is the volumetric transport rate per unit width, and s is the specific gravity of sediment, assumed to be 2.65. Combining (1) and (3) yields an equation for the total volumetric bed-load transport capacity,

$$Q_s = 11.2\frac{(\tau^* - \tau^*_c)^{4.5}}{\tau^{*3}}\ [(s\text{-}1)\ g\ D_{50}^{\ 3}]^{0.5}\ B, \tag{4}$$

where B is the channel width. Written in this form (4) illustrates that downstream variations in bed-load transport capacity are driven by the interaction of four main variables: τ^*, τ^*_c, D_{50}, and B. The effect of these interactions is evident in most streams and rivers, where, for example, downstream reductions in grain size are generally accompanied by increases in B and/or increases in the excess shear stress, τ^*- τ^*_c. The terms on the right-hand-side of (4) should therefore vary in ways that minimize the divergence in transport capacity, $\partial Q_s/\partial x$, otherwise aggradation or degradation will occur. It is also important to note that (4) includes two nonlinear terms, thus bed load transport computations are very sensitive to uncertainties in the individual variables. Among the variables in (4) we were most concerned with uncertainty in the estimates of τ^* and τ^*_c. Methods for estimating these parameters are discussed in detail below; however, it is worth noting that, in the case of gravel- or cobble-bed rivers, there is a strong theoretical and empirical basis for suggesting that the difference between τ^* and τ^*_c should be relatively small over the range of channel-forming flows [*Parker*; 1979; *Andrews*, 1984; *Pizzuto*, 1992; *Parker and Toro-Escobar*, 2002; *Pitlick and Cress*, 2002].

Uncertainties in measurements and calculations of shear stress in steep streams are due primarily to the fluid-mechanical effect of drag produced by large boulders or bank roughness

[*Bathurst et al.*, 1987; *Jarrett*, 1984; *Wiberg and Smith*, 1991; *Buffington and Montgomery*, 2001]. The reduction in effective (grain) stress due to drag on immobile boulders or increased bank roughness becomes increasingly important as channel gradient and relative roughness increase, as is the case here. However, our analysis does not include a correction for the effects of form drag [*Wiberg and Smith*, 1991]. We reasoned that existing data sets and measurements of bed load transport in laboratory flumes and steep streams have, to a varying extent, incorporated the effects of form drag in the measurements, without such a correction [*Shields*, 1936; *Aksoy*, 1973; *Bathurst et al.*, 1987; *Buffington and Montgomery*, 1997; *Shvidchenko and Pender*, 2000]. We therefore developed an empirical relation based on field data to estimate τ^*_c from reach-average slope, S, since this appears to be an important control [*Mueller et al.*, 2003]. Measurements of flow and bed load transport were obtained for 45 gravel-bed streams and rivers over a range in grain size and slope. For each data set, we formulated a relation between transport rate and dimensionless shear stress, and following the procedure of *Parker et al.* [1982], determined the reference τ^* associated with a small, non-zero transport rate; we equate this value with the critical τ^* [*Mueller et al.*, 2003]. The resulting estimates of τ^*_c were then correlated to reach-average slope, giving

$$\tau^*_c = 2.16\ S + 0.021, \tag{5}$$

with $r^2 = 0.70$ and $p \ll 0.001$. We then formulated an equivalent relation for the bankfull τ^* in Halfmoon Creek, using field measurements of bankfull hydraulic radius, reach-average slope and bed surface D_{50}, with one site excluded (SH9, Table 1). That relation is

$$\tau^*_b = 2.14\ S + 0.034, \tag{6}$$

with $r^2 = 0.91$ and $p \ll 0.001$.

Equations 5 and 6 were developed independently of each other, yet they are remarkably similar. The lines formed by (5) and (6) are almost parallel, indicating the difference between τ^*_b and τ^*_c is very nearly the same over the range of observed slopes.

Bed-load transport frequencies were estimated for each site using a regional flow duration curve based on a relation between bankfull discharge, Q_b, and drainage area, A. Additional assumptions regarding the at-a-site relation between τ^* and Q were necessary to calculate bed load transport rates for each flow category.

Estimates of Q_b were made for each site using measured values of bankfull channel width and hydraulic radius, R, and calculated values of bankfull velocity, U. The estimates of velocity were based on a flow resistance relation developed from current-meter measurements of discharge, taken over a range of flows (0.2-5.0 m^3/s) at 11 different sites. Velocity measurements were taken at closely spaced intervals across the channel (every 0.25 or 0.5 m) at 0.4 of the water depth. Four flows were measured at six of the main channel sites and two flows were measured at one main channel site and four of the tributary sites, yielding data for a range of depths and slopes. Values of the Darcy-Weisbach friction factor, f, were computed using

$$f = \frac{8gRS}{U^2}. \tag{7}$$

Values of f were then correlated to relative roughness, R/D_{84}, where D_{84} is the particle size for which 84% of the sediment is finer. A least-squares fit of these data gives the equation

$$\frac{1}{\sqrt{f}} = 2.26 \log\left(\frac{R}{D_{84}}\right) + 1.00 \tag{8}$$

with $r^2 = 0.70$ and $p \ll 0.001$, excluding outliers from one tributary site where very large boulders produced unusually high values of f. Over the range of observed flows the flow resistance

relation for Halfmoon Creek is very similar to a previous relation formulated by *Limerinos* [1970].

Values of the bankfull U for the 27 study sites were estimated from (7) and (8), then combined with the measured values of B and R to estimate Q_b. At some sites we observed that the cross-section based estimates of Q_b were anomalously high or low compared to nearby sites, primarily because of large differences in the estimated bankfull U. We therefore combined data from Halfmoon Creek with data from other gauged gravel-bed streams in Colorado [*Andrews,* 1984; *Torizzo and Pitlick,* 2004] to form a regional relation between Q_b and drainage area, A. Figure 3 shows a plot of the combined data, fit with the relation,

$$\hat{Q}_b = 0.35\, A^{0.725}, \tag{9}$$

with $r^2 = 0.88$ and $p \ll 0.001$, where $\hat{Q}_b$ is the estimated bankfull discharge used in subsequent calculations.

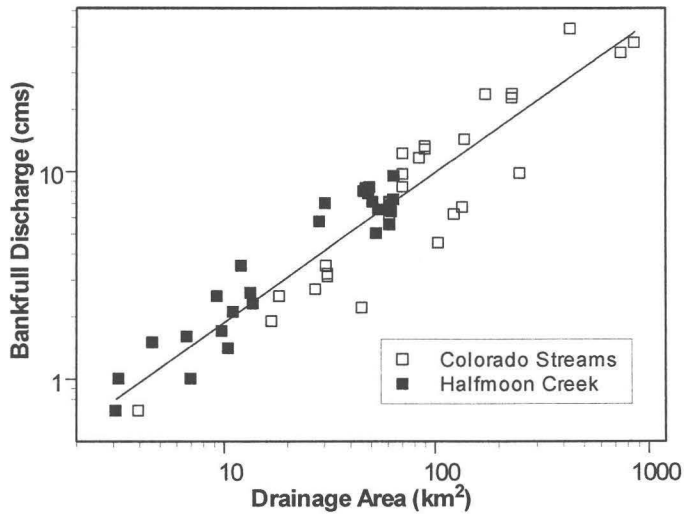

Figure 3. Regional relation between measured bankfull discharge, Q_b, and drainage area used to predict downstream changes in discharge for Halfmoon Creek.

A dimensionless flow duration curve was developed from the record of daily flows at the Halfmoon Creek gauge (record length = 55 yr). Daily discharge values were grouped into 40 intervals, and the interval values, Q_i, were normalized by the bankfull Q at the gauge (7 m^3/s). An exponential relation was fit to the upper end of the flow frequency distribution, giving the curve shown in Figure 4. This curve was then used to determine the frequency of individual discharges at each site, assuming that the form of the curve does not vary with drainage basin area (scale). This is a relatively good assumption for small basins in snowmelt-dominated regions because runoff intensity is driven primarily by surface-energy inputs (net radiation and sensible or latent heat transfer). It has been shown in a number of previous regional hydrology studies that the statistical distributions of flood peaks and daily flows in snowmelt dominated regions retain similar shapes across a wide range of drainage basin areas [*Pitlick*, 1994; *Gupta and Dawdy*, 1995; *Torrizo and Pitlick*, 2004].

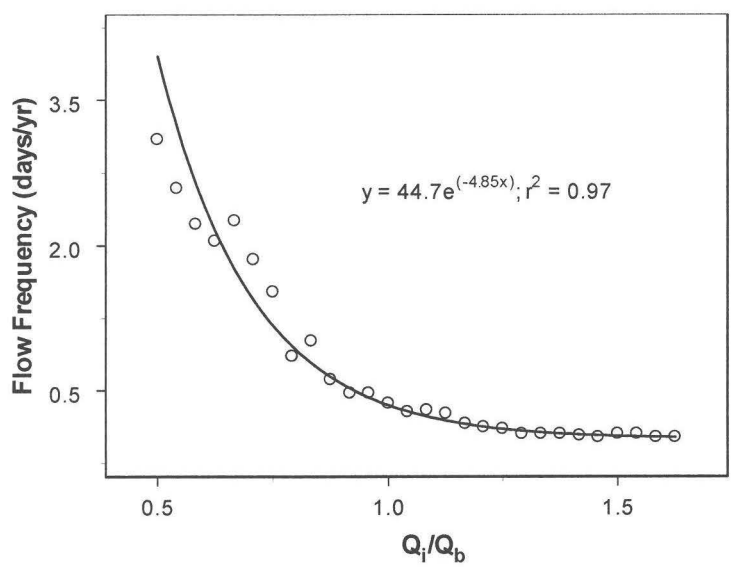

Figure 4. Relation between flow frequency and scaled discharge, Q_i/Q_b, for flows greater than 1/2 bankfull Q, as determined from the USGS gauge record.

Values of τ^* associated with the individual scaled discharges, $Q_i/\hat{Q}_b$, were estimated for each site using the function:

$$\tau^* = \hat{\tau}^*_b \left(\frac{Q_i}{\hat{Q}_b} \right)^{0.5}, \tag{10}$$

where $\hat{\tau}^*_b$ is the estimated bankfull τ^* for an individual site, found with (6), and the exponent 0.5 is the average of values obtained from measurements in 12 gravel-bed streams in Colorado [*Torizzo and Pitlick*, 2004]. Equation 10 has the desirable property that $\tau^* \rightarrow \hat{\tau}^*_b$ as $Q_i \rightarrow \hat{Q}_b$.

Annual bed load sediment yields were modeled for each site in the basin by determining Q_s for each value of $Q_i/\hat{Q}_b$; the values of Q_s were then multiplied by the flow frequency, and summed to estimate the annual yield. The choice of sediment transport equation may have some influence on the results; and in this study the Parker [1979] relation gave reasonable absolute quantities of sediment yield compared to the clear over-predictions of the Meyer-Peter Müller [1948] relation despite similar trends. Moreover, sediment transport equations based on excess shear stress are generally similar in form [*Gomez and Church*, 1989], thus the choice of equation is probably less important than calibrating it through selection of an appropriate critical shear stress [*Wilcock*, 2001; *Singer and Dunne*, 2004]. For this same reason, we tried to minimize uncertainty in our estimates of the bankfull τ^*, although this is clearly another important source of error. The potential error introduced by using a standardized flow-frequency relation is unknown, and there is some error associated with the field data; however, the uncertainty in these measurements and assumptions is probably small in comparison to the error in estimating τ^*_b and τ^*_c, hence much of our effort concentrated on developing relations for these parameters.

Results

Channel Slope

The longitudinal profile of Halfmoon Creek is concave overall (Fig. 5a), but marked by several distinct inflections associated primarily with glacial steps at tributary junctions [*MacGregor et al.*, 2000], but also debris fans and moraines. Headwater segments of the creek include short alluvial reaches separated by cascades, which appear as steps in the profile. The most abrupt change in gradient occurs in the area from 8-10 km where the three tributary forks come together. The creek then flows through an open alluvial valley and the profile is quite smooth. At a distance of ~5 km there is another small inflection in the profile where the creek is incising through glacial deposits and a debris fan produced by Elbert Creek. Below that point the valley again opens up, and for another 4 km, the slope decreases systematically downstream. In the lower 1 km of the study area, the creek is more confined by lateral moraines near the downstream extent of glaciation, resulting in another increase in slope.

In general, average gradients of tributary channel reaches are higher than main channel reaches (Table 1). A plot of average gradients of the study sites (Fig. 5b) shows that slopes of the tributary reaches generally exceed 0.025, whereas slopes of the main channel reaches are typically less than 0.025. As noted above, systematic decreases in slope are interrupted in several locations by glacial steps associated with tributary junctions and/or confinement between moraines and debris fans.

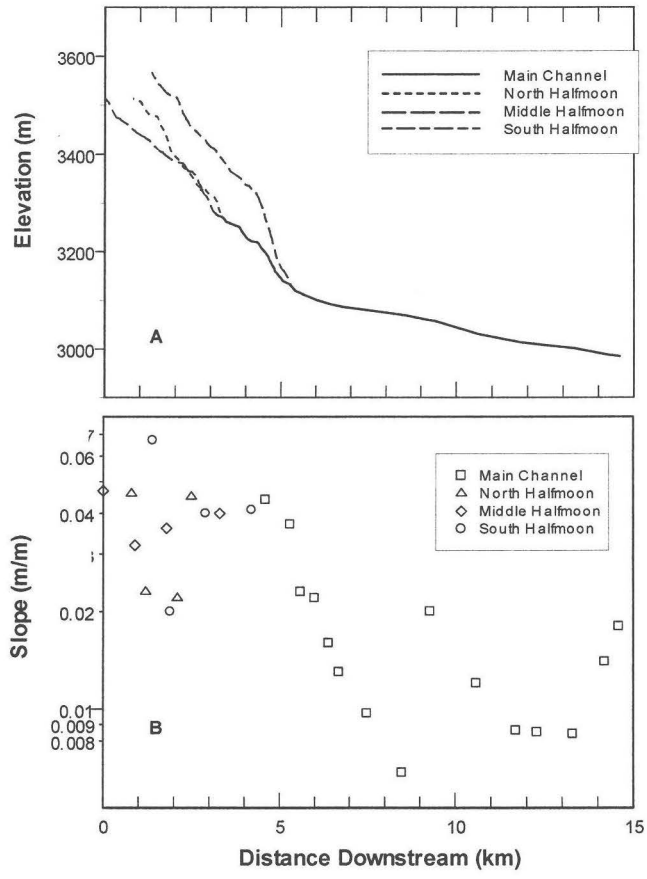

Figure 5. A) Longitudinal profile of Halfmoon Creek basin. B) Downstream changes in reach-averaged slope.

Bed Material

Median grain size of the surface bed material, D_{50}, in Halfmoon Creek varies from 0.086 to 0.030 m (Table 1, Fig. 6). Bed surface samples from the tributary channel reaches show little systematic variation or downstream reduction in grain size (Fig. 6), whereas samples from the main channel show systematic fining of the bed surface in two separate reaches, from 10-6 km and 5-1 km (Fig. 6). Downstream fining in these reaches mimics the rapid changes in slope noted in Figure 5, which reflect local flattening of the valley profile. *Rice and Church* [1998] describe such reaches as sedimentary links where changes in grain size are punctuated by

sediment inputs from tributaries or hillslopes. For purposes of comparison we fit the surface grain size data within these reaches with separate exponential functions:

$$D = D_o \, exp^{-\alpha (x - x_o)} \qquad (11)$$

where D_o is the grain size at the start of a fining sequence, x is the distance downstream from the start of a fining sequence, x_o, and α is a coefficient reflecting the rate of fining [*Sternberg,* 1875]. Table 2 compares regression estimates of the fining parameter α (km^{-1}) derived for surface and subsurface fining relations within links, as well as the basin as a whole. The results show that the surface D_{50} within the two main-stem sedimentary links fines very rapidly over short distances ($\alpha = 0.33$ and 0.11, respectively, $p < 0.01$, Table 2). The estimates of α for these two links are high, but comparable to sedimentary links in active gravel-bed rivers where localized sediment supply and effects of past glaciation sometimes force abrupt transitions in valley morphology and slope [*Ferguson and Ashworth*, 1991; *Knighton*, 1998; *Morris and Williams*, 1999; *Rice*, 1999]. The influence of irregular transitions in slope is especially pronounced in the tributary valleys and suggests that fining trends may not be resolved by the sampling interval ($\approx$ 1 km). In contrast to the individual links, the basin-wide trend in surface D_{50} in Halfmoon Creek is weak ($\alpha = 0.012$), and the relation is not statistically significant ($p = 0.23$).

Downstream variations in the size distribution of the subsurface sediment are not nearly as pronounced as the surface sediment discussed above. The median grain size of the subsurface sediment, D_{50s}, ranges from 0.017 to 0.036 m (Table 1). A plot of the downstream trends in D_{50s} (Fig. 6) displays minor inflections at some of the same points noted above. However, fining below these points is not strong, and values of α for the subsurface D_{50} within individual links are not statistically significant, nor is the coefficient for the basin-wide relation ($p > 0.2$, Table 2).

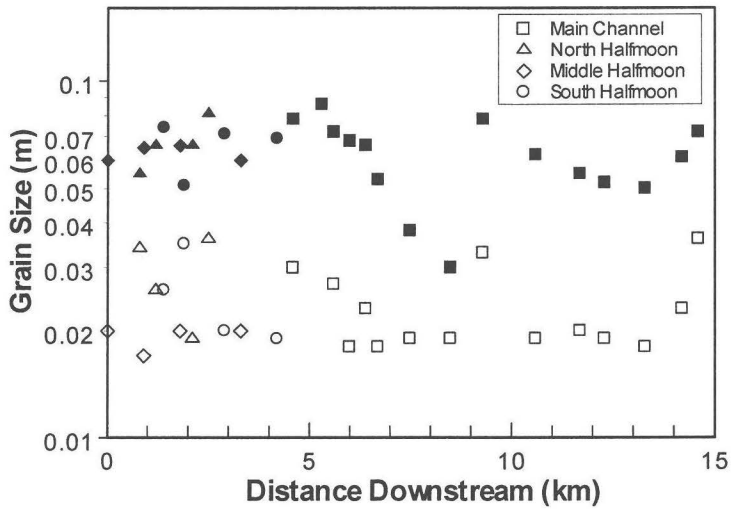

Figure 6. Downstream changes in the median grain size of the surface and subsurface bed material. Black points are surface D_{50} and transparent points are subsurface D_{50}.

Table 2. Comparison of surface and substrate fining relations for Halfmoon Creek. The fining coefficient, α, was determined from linear regression of ln-transformed values of grain size versus distance; n is the number of values, r^2 is the coefficient of determination, F is the value of the F distribution, and p is the significance probability.

Surface	**n**	**α**	**r^2**	**F**	**p**
10-6 km	7	0.33	0.97	187	«0.001
5-1 km	5	0.11	0.94	44	0.007
Overall	27	0.012	0.06	1.4	0.23
Substrate					
10-6 km	6	0.076	0.25	1.3	0.32
5-1 km	5	0.201	0.43	2.3	0.23
Overall	26	0.014	0.04	1.1	0.31

Downstream fining within the sedimentary links is likely caused by a combination of size-selective transport, and to a lesser extent, abrasion. Peak snowmelt flows in Halfmoon Creek rarely exceed the bankfull discharge by more than a factor of two, thus the condition of partial transport [*Wilcock and McArdell*, 1993] prevails for most high flows. Results presented below show that the difference between coarse and intermediate size fractions varies with slope and shear stress. Fining within the sedimentary links may, therefore, be interpreted as a textural response that enhances transport through reaches with rapidly decreasing shear stress. While changes in supply may also alter trends in surface grain size [e.g. *Dietrich et al.*, 1989], channel reaches within the fining links are typically disconnected from hillslopes. Abrasion may also contribute to fining within links, as much of the basin is underlain by thinly-foliated metamorphic rocks which vary in their resistance to weathering. Sediment worn down by abrasion may thus be re-supplied by localized erosion of moraines and debris fans.

Figure 7 shows grain size distribution of all sediment samples, differentiated by reach-average gradient. We differentiated the samples to examine connections between channel gradient and size-separation of the individual fractions. The split was made at $S = 2.5\%$ because it is approximately at this point where we start to see major differences in hydraulics, and because this split results in roughly equal samples sizes. Close inspection of the data in Figure 7 suggests that the differences in sediment samples from high- and low-gradient reaches are mostly the result of variations in the coarse size fractions. To evaluate the significance of these variations we performed a series of *t* tests comparing differences among the individual grain size parameters (D_{84}, D_{50}, and D_{16}). The results are presented in Figure 8, which shows a series of box plots comparing parameters of surface and subsurface samples, differentiated by gradient. The first panel (Fig. 8a) indicates that the average surface D_{84} of high- and low-gradient reaches

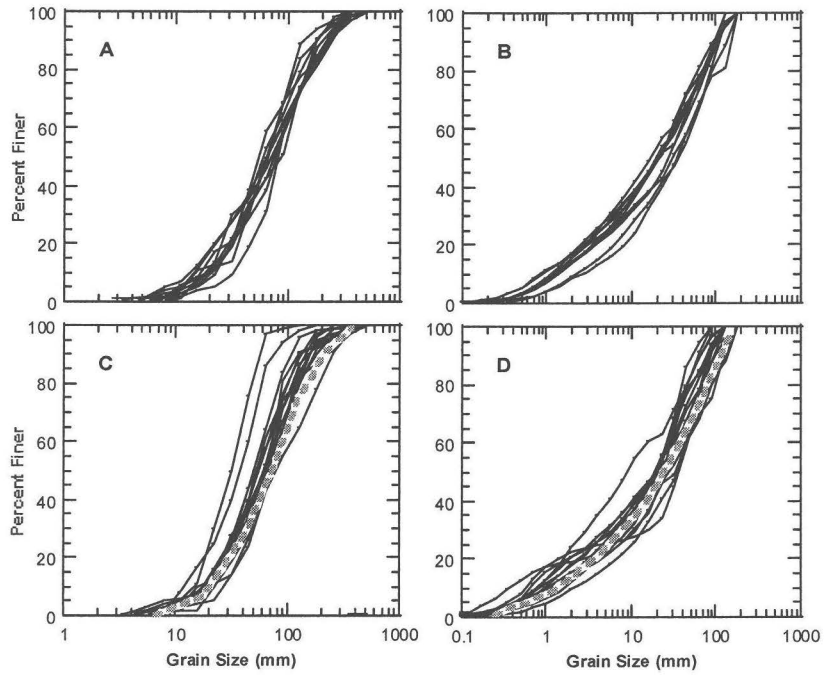

Figure 7. Full grain size distribution of all sediment samples. A) and B) represent surface and subsurface samples taken in high-gradient reaches ($S > 2.5\%$), respectively. C) and D) represent surface and subsurface samples taken in low-gradient reaches ($S < 2.5\%$), respectively. The dashed grey lines represent the average grain size distributions of the high-gradient reaches.

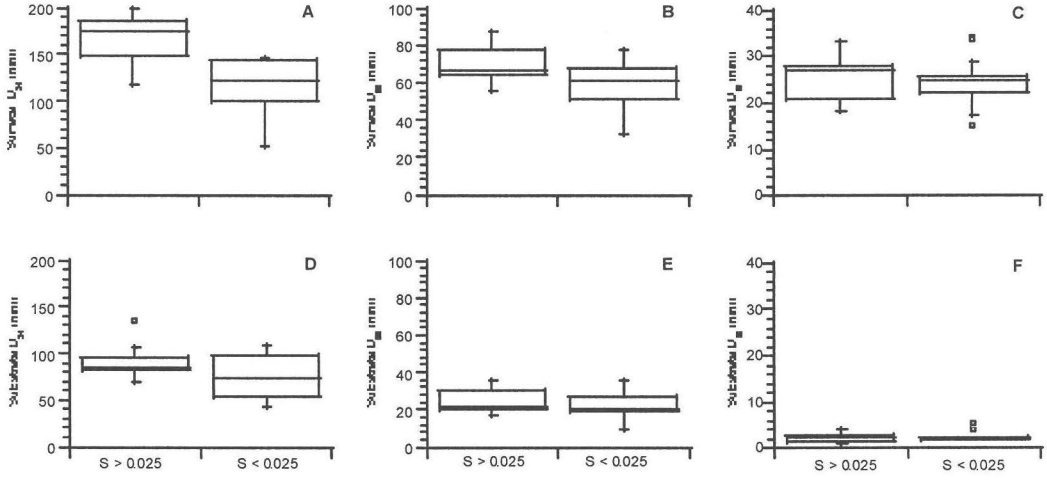

Figure 8. Box plots illustrating the median, quartiles, minimum, maximum, and outliers in A) surface D_{84}, B) surface D_{50}, C) surface D_{16}, D) subsurface D_{84}, E) subsurface D_{50}, and F) subsurface D_{16} between high- and low-gradient reaches.

differs by 46 mm (38%), which is statistically significant ($t = 3.56$, $p = 0.0015$). The second panel (Fig. 8b) shows that there is also a difference in the average surface D_{50} of high- and low-gradient reaches, but the separation is much less (11 mm or 18%) and the statistical significance of the difference is lower ($t = 2.39$, $p = 0.025$). The third panel (Fig. 8c) shows that there is no significant difference in the average D_{16} of the surface sediment ($t = 0.79$, $p = 0.44$). Figures 8d-8f compare parameters of the subsurface size distributions. Figure 8d indicates that the average subsurface D_{84} of high- and low-gradient reaches differs by 17 mm (23%), which is marginally significant ($t = 2.06$, $p = 0.05$). The final two panels (Figs. 8e and 8f) comparing values of the subsurface D_{50} and D_{16} show that there is no significant difference ($t = 0.52$, $p = .61$; $t = 1.52$, $p = 0.14$) between samples taken in high and low-gradient reaches.

The results presented above indicate that the distinction between surface and subsurface sediment in Halfmoon Creek is pronounced, particularly in the steeper reaches. The size distribution of the subsurface sediment is relatively consistent throughout the basin, but there is a clear tendency for the surface layer to become relatively coarser in steeper reaches. Supposing the size distribution of the bed load is similar to the subsurface sediment, as proposed by *Parker et al.*, [1982], we postulate that the effects of grain hiding and exposure associated with changes in bed surface texture should, to a certain extent, counter changes in the mobility of the bed load as it moves through steeper or flatter reaches. These interactions are examined in the next section, which describes the downstream trends in channel geometry and the linkages between flow, boundary shear stress and bed sediment properties.

Channel Morphology and Hydraulics

The average bankfull width, *B*, of Halfmoon Creek increases from about 3 m in headwater reaches to about 11 m in lower reaches (Table 1, Fig. 9a). Over the same distance, the average bankfull hydraulic radius, *R*, increases from about 0.3 to 0.5 m (Table 1, Fig. 9b). The data follow more irregular trends in the middle of the study area where the slope and grain size are changing rapidly. Between sites 20 and 13, for example, there is a decrease in bankfull *R* and an increase in bankfull *B* coinciding with fining through the upper sedimentary link and a transition in planform from straight plane-bed reaches to more sinuous pool-riffle reaches. This trend reverses at site 10 where the Elbert Creek fan starts to confine the valley and supply coarse bed material. The trends of decreasing bankfull *R* and increasing bankfull *B* are crudely repeated through the lower sedimentary link (sites 10-4), again without much change in water discharge.

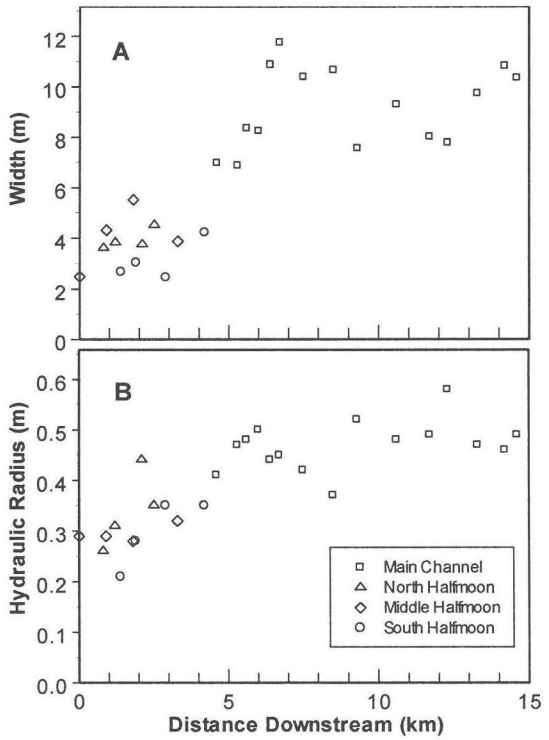

Figure 9. Downstream trends in A) width and B) hydraulic radius.

Downstream trends in the bankfull shear stress, τ_b, and the bankfull Shield's stress, τ^*_b, are plotted in Figure 10. Coding the individual values by average slope, S, shows that gradient has a disproportionate influence on local estimates of both τ_b and τ^*_b; in general, the highest values of τ_b and τ^*_b occur in reaches with $S > 2.5\%$.

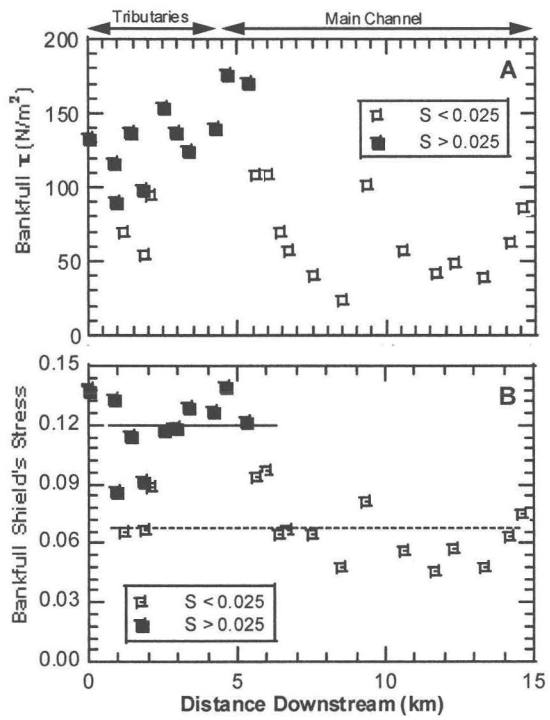

Figure 10. Downstream trends in A) bankfull shear stress, τ, and B) bankfull dimensionless shear stress, τ^* with data stratified by slope.

The strong association between gradient and shear stress suggests that both variables interact with mobile fractions of the bed sediment to give the textural trends discussed earlier. These interactions are examined in Figure 11, which plots four separate relations between the bankfull shear stress and sediment size, measured in terms of two parameters (D_{84} and D_{50}) of the surface and subsurface sediment. Among the four parameters, the surface D_{84} is most highly

correlated to τ_b (Table 3); the subsurface D_{84} and the surface D_{50} are likewise correlated to τ_b, although the trends are not quite as strong as in the previous case (Table 3). Note that the relations for the subsurface D_{84} and surface D_{50} nearly coincide, suggesting that, for a given τ_b, the coarser fractions of the bed substrate actively exchange with average-sized particles on the bed surface. The last relation shown at the bottom of Figure 11 indicates that the subsurface D_{50} is weakly correlated to τ_b; the significance probability of this relation ($p = 0.12$; Table 3) suggests that the slope of the line is not significantly different from zero, or in other words, the subsurface D_{50} is essentially constant in relation to the bankfull shear stress.

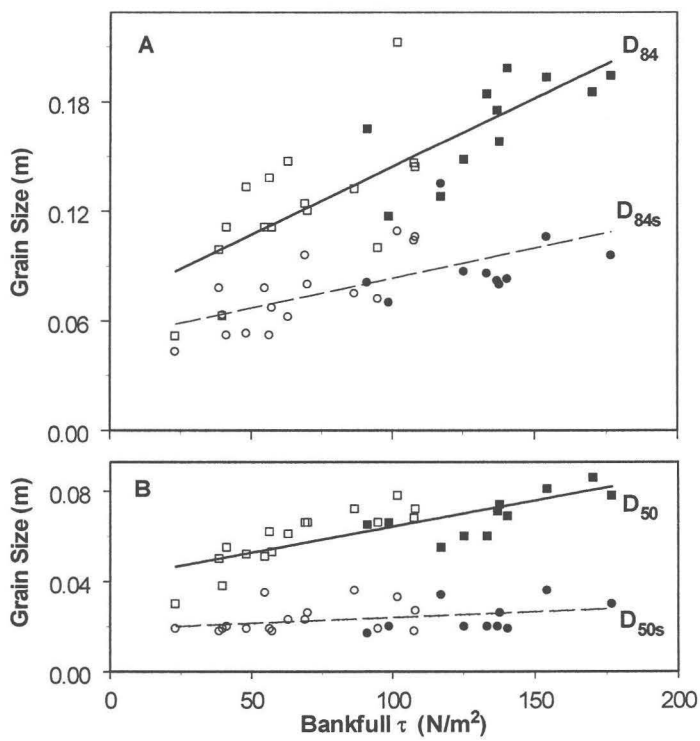

Figure 11. Relation between bankfull shear stress, τ, and A) surface and subsurface D_{84} and B) surface and subsurface D_{50}. Black and white symbols represent high- and low-gradient reaches, respectively.

Table 3. Linear regression statistics of grain size versus bankfull shear stress for the surface and subsurface D_{50} and D_{84} for data in Figure 9; a is the slope and b is the intercept of the of the regression equations, r^2 is the coefficient of determination, and p is the significance probability.

Grain Size	a	b	r^2	p
D_{50}	0.00023	0.041	0.62	«0.001
D_{50s}	0.00005	0.019	0.11	0.11
D_{84}	0.00074	0.070	0.64	«0.001
D_{84s}	0.00033	0.051	0.42	0.0004

The relations shown in Figure 11 suggest that the separation between intermediate and coarse size fractions becomes increasing large as shear stress increases. These differences are particularly evident in comparing the substrate D_{50} with the size parameters of the bed surface: in reaches with relatively low shear stress ($\tau < 50$ N/m^2), the substrate D_{50} differs from the surface D_{50} by a factor of about 2, whereas in reaches with high shear stress ($\tau > 150$ N/m^2), the difference exceeds a factor of 3. Thus, while it might appear that the steeper reaches of Halfmoon Creek have a high capacity for carrying fine-medium gravel, the relative mobility of these sizes is affected by the presence of very coarse gravels and cobbles on the bed surface. The proportion of surface-layer grains moving during high flows is probably never very large, however, it was evident during field sampling that many of the coarsest grains on the bed were loose and likely mobile at high flow. The substrate exists presumably because there is exchange between the sediment in motion and the sediment stored in the bed. Exchange between the bed load and the substrate cannot occur unless there is some movement of the surface layer grains, and as suggested below, this appears to occur at all of the study sites at least a few days per year.

Sediment Transport Model

Modeled annual sediment yields are plotted in relation to bankfull discharge in Figure 12. Although the modeled values exhibit appreciable scatter, the overall trend is consistent with the expectation that sediment-transport capacity should increase monotonically downstream. These data were fit with a power law relation:

$$Q_s = 23.2\ \hat{Q}_b^{\,0.96}, \tag{12}$$

with $r^2 = 0.92$ and $p \ll 0.001$, where Q_s is in metric tons per year. The observed trend in Q_s

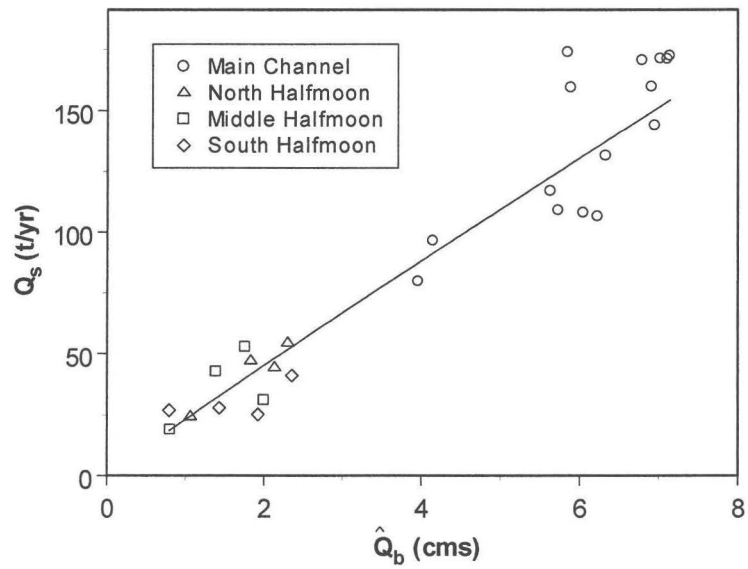

Figure 12. Modeled annual sediment yield, Q_s, plotted versus estimated bankfull discharge, $\hat{Q}_b$.

reflects potentially complex interactions among the variables in (4). In high-gradient reaches with coarse bed material, instantaneous unit transport rates (load per unit width per unit time) tend to be very high; however, since it requires a relatively high discharge to exceed the estimated threshold Shield's stress, $\hat{\tau}^*_c$, bed load transport in these reaches occurs perhaps only a few days per year, and as a result, the average annual load is not particularly high. In low-

gradient reaches with finer bed material, $\hat{\tau}^*_c$ is exceeded more often and sediment is moved perhaps many days per year. The effects of low shear stress and small grain size in low-gradient reaches are thus more than made up for by an increase in the frequency of bed load transport, resulting in relatively high annual loads.

Interestingly, the estimated annual sediment yield increases nearly linearly with discharge (Fig. 12), suggesting a simple scaling of sediment yield as discharge increases downstream. The frequency of sediment transporting flows ranges from just under 3 days/yr at the steepest site, to nearly 18 days/yr at the lowest gradient site. The variations in flow frequency are primarily related to the non-linear relation between Q_i/Q_b and τ^* because at steeper sites where $\hat{\tau}^*_c$ is higher, a slightly higher unit discharge is necessary to initiate motion. Despite variations in grain size, width, flow frequency, and channel gradient, the modeled sediment transport relations suggest that these reaches convey sediment from upstream in some near-equilibrium state where the total bed load increases downstream in proportion to the water discharge.

Discussion

The principal goals of this research were to investigate interactions among channel- and bed-material properties within a headwater stream, and then use those properties to model basin-wide trends in annual bed-load sediment yield. The results indicate that in spite of significant variations in channel morphology and sediment texture, the modeled sediment yields define a clear trend, with bed-load transport capacity increasing nearly linearly with discharge. A number of empirical relations and assumptions went into the development of the model, thus the question arises: Is it possible to validate the results in any way? Since we did not measure bed

load at any of the sites, we cannot assess the accuracy of the estimates directly, and as far as we know there isn't a comparable data set that allows us to do this. However, a reasonable comparison can be made using downstream relations for suspended sediment, for which there are several examples. *Emmett* [1975] measured suspended sediment at 20 stations in the upper Salmon River drainage basin in Idaho. He used the at-a-station relations for suspended sediment to formulate a downstream relation for load as a function of bankfull discharge,

$$Q_{ss} = 0.89\ Q_b^{0.75} \tag{13}$$

where Q_{ss} is the suspended sediment load, in tons per day, and Q_b is the bankfull discharge, in cubic feet per second. *Leopold and Maddock* [1953] developed a set of flow and suspended-sediment relations for 20 rivers in the western USA, and suggested that the exponent in (13) should have a value of ≈ 0.80. The close agreement between the exponents in these two previous relations and our relation (0.96) is encouraging, however, we recognize the potential problems in comparing bed-load and suspended-load transport as parallel processes.

Variations in bed sediment texture have been described in the literature as a response to changes in sediment supply and/or transport capacity. Results from a number of laboratory experiments generally indicate that the bed surface becomes finer as the sediment supply increases [*Dietrich et al.*, 1989; *Lisle et al.*, 1993; *Buffington and Montgomery*, 1999b], or coarser as the supply decreases [see *Lisle and Church*, 2002, and references cited therein]. The surface-based transport model of *Parker* [1990] likewise suggests that the surface texture will become finer or coarser as necessary to balance the sediment supply. However, *Wilcock et al.* [2001] conducted a series of experiments in a sediment-recirculating flume, and showed that, if anything, the bed surface becomes coarser as discharge and transport rate are increased. The contrast in results stems from differences in experimental design, which have an important effect

on the size distribution of the sediment supplied. In typical sediment feed experiments the same sediment is used throughout the experiment and only the supply rate is varied with the discharge; the size distribution is thus kept constant. As the supply rate is increased, the bed is observed to become finer, which must happen in order to balance transport rates of individual size fractions. In recirculating experiments the size distribution of the input is governed by the output, which evolves over time because of changes in bed surface texture. Natural streams are hybrids of these two systems. Furthermore, alluvial channels are unlike flumes in that width is adjustable. Adjustments in width provide another mechanism for a channel to respond to changes in sediment load, although perhaps not over the same time-scale as changes in bed texture.

Given these observations, we want to avoid reading too much into the trends in surface and substrate grain size distributions in Halfmoon Creek. The data clearly show that there is not much difference in the D_{50} of the substrate throughout the basin, suggesting that the sediment supplied from hillslopes and channel-bed sources has a similar size distribution. However, the coarser fractions of the substrate and the bed surface clearly vary with the local shear stress. It appears that the shear stress and grain size vary jointly in order to balance the frequency and intensity of bed load transport; adjacent reaches with very different slopes and grain sizes appear to carry roughly the same load. Given the overall stability of the basin, conditions in Halfmoon Creek are probably more analogous to a sediment-recirculating system than a sediment-feed system. Hillslopes are not strongly coupled to the channel, thus the "feed" to any particular reach is determined largely by the output from the reaches upstream. Glaciation, however, has left a strong imprint on the landscape, producing reaches with locally high slope and shear stress. The coarse texture of the bed sediment in the steep reaches could, therefore, be viewed as a mobile bed response to changes in shear stress and the size distribution of the supply. Further

study of these interactions is needed to clarify the relation between discharge, sediment supply and bed sediment texture, as it does not appear that such a relation is easily deduced from field measurements of a few variables.

Our results indicate that Halfmoon Creek and it's tributaries appear to function more or less as an alluvial stream system, even at slopes as high as 6%. However, the analysis does not provide unambiguous evidence, one way or the other, that Halfmoon Creek transports sediment at capacity. *Gilbert* [1914, p. 35] defined sediment-transport capacity as "the maximum load a stream can carry", presumably for a given condition; load was defined as the quantity (mass or volume) of sediment carried through a cross section per unit time. In addition to discharge and slope, transport capacity varies with grain size [*Gilbert*, 1914; 1917; *Mackin*,1948; *Rubey*, 1952]. The load can be measured, whereas the transport capacity can only be inferred. Typically, those inferences are made on the basis of comparisons between measured and predicted transport rates. Since we did not measure transport rates we cannot, in the strictest sense, make that comparison. For the same reason, however, there is a danger in basing estimates of transport capacity on flow strength (or a suitable surrogate, such as drainage area times slope), because without reference to a threshold flow or a characteristic grain size, there is no clear basis for comparison there either. The processes that govern erosion and sediment transport in mountain streams can, therefore, be more fully understood if we begin to couple high-quality transport measurements, obtained over a full range of flows, with basin-wide measurements of channel morphology and grain size.

Conclusion

The channel and bed-material properties of Halfmoon Creek, a headwater stream in the Sawatch Range of the southern Rocky Mountains, reflect a combination of free and forced responses to changes in valley slope, water discharge and sediment supply. Short steep reaches

that begin in the vicinity of glacial steps, moraines or debris fans are characterized by very coarse bed material and high shear stress. However, sediment within these reaches appears to be mobile for at least a few days per year. Selective transport of finer sediment (medium-fine gravel) through the steep reaches produces rapid downstream fining and less of a contrast between the bed surface layer (armor) and the substrate sediment. We postulate that, with a reduction in bed armoring, the overall mobility of the bed load is enhanced, which increases the frequency of transport, and thereby maintains continuity through reaches with declining shear stress. Sediment-transport modeling suggests that forced changes in channel gradient are offset by changes in channel width, depth, and bed surface texture to maintain a rough balance between the intensity and frequency of bed load transport. The modeled sediment yields indicate that the bed load transport capacity of the creek increases systematically downstream with discharge, similar to previous empirical relations formulated for suspended sediment.

Acknowledgements

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